Robustness of The Nonlinear Climate Response to ENSO's Extreme Phases

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ABSTRACT

Analysis of a suite of atmospheric general circulation model experiments for 1950-94 shows that both the tropical and the extratropical wintertime climate respond nonlinearly with respect to opposite phases of ENSO. Such behavior is found to be reproducible among four different GCMs studied, confirming that a similar behaviour witnessed in observations is indeed symptomatic of nonlinearity rather than sampling error.

Nonlinearity in the tropical Pacific rainfall response is related to an SST threshold for convection that leads to saturation at modestly cold SST forcing, but a linear increase for warmer SST forcing. A spatial shift in the rainfall response is also a feature of the various GCMs' nonlinear behavior, that is accentuated by the large zonal gradient of climatological SSTs across the equatorial Pacific and the fact that convection responds to the total rather than the anomalous SST.

Regarding upper tropospheric teleconnection responses over the Pacific-North American region, nonlinearity exists in both the spatial phase and the strength of the midlatitude response. The four GCMs are found to be unanimous in having their teleconnection patterns shifted eastward during warmer tropical Pacific SSTs relative to colder SSTs. Furthermore, the increase in amplitude of the extratropical atmospheric response for a standardized increase in warm tropical Pacific SSTs is roughly double that for an identical increase in cold SST forcing.

Our further analysis of model simulations reveals that nonlinearity in climate responses emerges mainly for stronger ENSO events, and a predominately linear response is found for weaker tropical Pacific SST forcing. In particular, climate simulations using both realistic and idealized SSTs indicate that tropical Pacific SST anomalies greater than one standard deviation of the interannual variation are required for initiating an appeciable nonlinear climate response.

1. Introduction

In a recent workshop that reviewed causes and consequences of cold events (Glantz 1999), it was pointed out that conclusions on cold event impacts have largely been drawn from studies on warm events. Yet, recent results suggest that the global climate signal associated with El Nino-Southern Oscillation (ENSO) is not simply linear with respect to ENSO's warm and cold phases. Among the recommendations emerging from the workshop was the need to understand those situations and locations under which ENSO impacts on climate and society differ from the typical assumption of linearity and symmetry.

Distinguishing the true nonlinear signals from the climate noise is difficult from observational analysis alone, and independent information gathered from numerical models is essential. Some analysis on this problem has been made using general circulation model (GCM) sensitivity experiments, however results have been inconsistent among the different GCMs employed (e.g., Cubash 1985; Hoerling et al. 1996; Gershenov and Barnett 1998). An additional issue of importance is the fact that the nonlinearity in atmospheric responses will itself depend on the tropical Pacific sea surface temperatures (SSTs) (Kumar and Hoerling 1998). To date, however, no systematic GCM analysis has been undertaken that examines the nonlinearity comprehensively for the requisite broad spectrum of SST forcings.

Given that the skill of seasonal climate forecasts hinges on predicting the ENSO signal accurately, the practical benefits of understanding the higher-order aspects of that signal related to nonlinearity may be significant. The extent to which improvements in seasonal forecast skill can be expected through incorporating the effects of warm and cold events separately will ultimately depend on clarifying the degree of nonlinearity of atmospheric responses, and identifying the SST states conducive for such behavior.

In this study on the problem of nonlinear climate responses to ENSO extreme phases, we will present results based on realistic and idealized SST anomaly experiments. To insure that our analysis is not unduly effected by biases in any particular GCM, four different atmospheric models, as described in section 2, are subjected to identical monthly SST evolutions spanning the last half century, and for each model an ensemble of runs is performed. Each ensemble is subjected to identical statistical analysis in order to identify the robust features of the seasonally averaged atmospheric response to warm and cold events separately, and results based on realistic SST scenario experiments are presented in section 3.

Our analysis of those results focuses on two properties of the wintertime response to ENSO that have been previously identified as features indicative of nonlinearity. First, the tropical Pacific rainfall response, for which an inherent nonlinearity with underlying sea surface temperatures has been long known, is examined. We explore especially the robustness of the saturation effect on tropical Pacific rainfall responses during cold ENSO states as suggested previously in Kumar and Hoerling (1998), and the phase shift of equatorial Pacific rainfall anomalies for warm events relative to cold events as shown in the composites of Hoerling et al. (1997).

Second, the upper tropospheric teleconnection patterns over the Pacific-North American (PNA) region, whose prominent phase shift with respect to warm and cold events has been argued to explain nonlinearities in the North American land temperature and rainfall anomalies (e.g., Montroy et al. 1998; Gershenov and Barnett 1998), are analyzed. We will focus on the amplitude of those teleconnections as a function of ENSO's phase, with the understanding that the strength of the signal (relative to the climate noise) determines to large degree the potential predictability. On this matter, we seek to reconcile differences among prior GCM results that on the one hand indicate the cold event teleconnection signal to be negligable and thus possess low predictability (Cubash 1985), whereas on the other hand Hoerling et al. (1997) and Molteni et al. (1993) find the cold event signal to be appreciable and comparable to its warm event counterpart.

The longitudinal phase of the simulated teleconnections with respect to cold and warm events is also diagnosed. Our intent here is to provide an interpretation of Hoerling et al's observational result that the warm event 500 mb height anomaly composite has centers-of-action over the PNA region shifted 30 degrees longitude east of their cold event counterparts.

Finally, section 3 also presents the results of a suite of idealized SST anomaly experiments using one of the aforementioned atmospheric GCMs. These further examine the extent to which nonlinearity is a function of the amplitude of the SST forcing. A summary with concluding remarks appears in section 4.

2. Data sets and methods

a. Observational data

Global sea surface temperatures are analyzed for the period during 1948-99. The SST data set is identical to that described previously in Hoerling et al. (1997) except that it has been updated to include analyses through February 1999.

The monthly SST data is subjected to an empirical orthogonal function (EOF) analysis in order to derive an objective index of the state of the tropical Pacific Ocean. The EOF analysis is of the co-variance matrix that consists of 614 monthly time samples of anomalous SSTs from January 1948-February 1999. The analysis is conducted over the region 30°N-30°S, 120°E-60°W, and the first EOF (E1) shown in Fig. 1 explains 45% of the total SST variance over this domain. The principal component (PC) time series is shown in Fig. 1a, and positive departures are associated with a "warm phase" of the eigenvector in Fig. 1b.

The observed cold season (December, January, February, and March) anomalies in 500 mb height, rainfall, and surface temperature associated with ENSO are calculated by regressing the particular variable upon the SST index as given by the time series in Fig. 1. The National Center for Environmental Prediction's (NCEP) monthly mean 500 mb heights spanning the period January 1948-February 1999 is available on a 2.5° grid for the region 20°N-90°N. Global monthly mean surface temperature and rainfall for the same period are derived from the Global Historical Climate Network (GHCN) data, and these have been gridded on a regular 2.5° grid. A more detailed spatial analysis of the tropical rainfall signals, albeit based on a shorter record, is derived from polar orbiting satellite measurements of outgoing longwave radiation (OLR) spanning the period March 1974 to February 1999 (Liebmann and Smith 1996). Seasonally averaged rainfall anomalies are estimated from the OLR data using the empirical relation of Arkin and Meisner (1987).

b. Atmospheric general circulation model experiments

Monthly mean global SSTs for the 1950-94 period are imposed as evolving lower boundary forcing for a suite of atmospheric GCM simulations. Four different models, whose characteristics are summarized in Table 1, are employed, and multiple realizations of the 45-year period are generated for each GCM. The ensemble members of a particular model differ from each other only in the specification of the atmospheric initial conditions, and each realization experiences the same evolving global SST boundary conditions. The model runs all begin on January 1950, using initial atmospheric states selected from control integrations of the respective GCM.

All the models employed are global spectral, and use sigma coordinates for vertical discretization. Table 1 summarizes the spectral truncations of the GCMs, and the effective horizontal resolution for all the models is roughly 3° lat/lon. Different parameterizations of moist physics are used in the various GCMs, and here we highlight their deep convection schemes. NCAR's CCM3 uses a mass flux method as described in Hack (1994) together with the deep cumulus formalism of Zhang and McFarlane (1995). The ECHAM-3 model also uses a mass flux scheme, though following the formalism of Tiedtke (1989) and as described further by Roeckner et al. (1992). The GFDL model uses moist adiabatic adjustment, whereas the MRF climate model uses a modified version of the Kuo scheme. The reader is referred to the cited references in Table 1 for further details of the dynamical formulations and parameterized physics employed.

An additional set of GCM simulations are diagnosed that employ idealized SST forcing, and these runs utilize the MRF climate model. In these runs, the first EOF of tropical Pacific SSTs (see Fig. 1) is prescribed as the only spatial structure of interannual SST variability. This pattern undergoes a monthly evolution in amplitude and phase from January 1965-December 1989 that is given by the PC time series of Fig. 1a. A second set of parallel runs is performed in which the opposite phase of the PC time series of Fig. 1a is used to define the monthly evolution of the E1 tropical Pacific SSTs. Thus, analogous to the idealized experiments of Hoerling et al. (1997), this suite facilitates comparison of warm events that are now precisely the inverse of cold event counterparts. These new experiments, however, permit analysis of nonlinearity of atmospheric responses for various amplitudes of the SST forcing rather than for only a single amplitude forcing as in Hoerling et al.. For each set of experiments, a nine member ensemble has been generated using the method described previously.

c. One-sided linear regression method

One-sided regressions are performed separately for the warm and the cold phases of E1. The procedure involves calculating the linear relation between all occurrences of one sign of the SST index and the corresponding predictand. Anomalies used in the analysis are calculated with respect to means based on samples associated with a single sign of the SST index. The one-sided regressions are estimated from 26 years of data (half the record) for the observational analysis, and 22 years of ensemble averaged data for the GCM analysis The regressions thus indicate the change of a particular atmospheric variable for a unit change of warm (or cold) tropical Pacific SST, and both warm and cold one-sided regression results in section 3 are presented for a one standardized departure of the E1 SST index.

In the method of one-sided regression as implemented herein, the regression lines cut through an "origin" that is defined by the point corresponding to a composite warm (cold) event anomaly itself. Figure 2 shows schematically the frame of reference in which the regressions are calculated. The warm and cold event composites define the origin of the two embedded axes in Fig. 2, and in the current analysis these are equal and opposite to each other since each is based on half the data record. Figure 3 illustrates the composite for warm SST conditions, using 500 mb height and tropical rainfall, for the observations (top panel), and for the grand ensemble average of all four GCMS (bottom panel). A summation of the one-sided standardized regression with its corresponding composite map estimates the atmospheric anomaly related to a roughly two-standardized SST anomaly. The one-sided regression, however, contain all the information about potential nonlinearities of the atmospheric response with respect to ENSO's phase.

3. Results

a. Analysis of observations

One-sided regressions reveal substantial asymmetries in both tropical and extratropical Pacific climate anomalies associated with ENSO's opposite phases (Fig. 4). The change in tropical Pacific rainfall is both more intense and more expansive for a unit change in warm SST forcing than it is for a unit change in cold SST forcing. For warm SSTs, the change in rainfall possess a di-polar structure (Fig. 3a) that resembles the well-known composite warm event di-pole pattern itself, but with centers shifted farther toward the east equatorial Pacific. In contrast, the rainfall regression for cold events is a negative monopole, whose center slightly west of the dateline is co-located with the composite cold event signal (opposite phase of Fig. 2a) suggesting that the rainfall pattern for progressively colder SSTs undergoes little change.

Also noteworthy in Fig. 4 is that the observed changes in 500 mb height for a unit increase in warm SSTs are as much as double those related to a unit increase in cold SST forcing. For the PNA-region bounded by 150E-60W, 20N-70N, the root mean square (rms) height regression is 23 m for a one standardized change in warm SSTs, and 15 m for an equivalent change in cold SSTs. This comparative weakness of the cold regression height amplitudes is qualitatively consistent with the weakness of the rainfall anomalies during cold SST states.

A further distinction of the height regressions is their spatial phase shift, and the correlation between the warm and cold 500 mb height patterns of Fig. 4 is only -0.4 over the PNA region. This phase shift in the upper tropospheric circulations is widely believed to explain the phase shift in North American surface temperature anomalies for warm relative to cold events as illustrated in Fig. 5. Whereas the continental surface temperature regressions have comparable strength, a tendency toward below normal temperatures for colder tropical Pacific SSTs occurs over Alaska and the Yukon Territory, whereas a tendency toward above normal temperatures for warmer tropical Pacific SSTs occurs over the Canadian Plains and the Great Lakes region.

North American wintertime precipitation (Fig. 6) appears to respond more linearly to ENSO's extreme phases than does temperature, and only along the Pacific West Coast is there appeciable asymmetry. Note especially the increase in rainfall from coastal British Columbia southward to Baja for warm SSTs, suggesting that the entire US West Coast is prone to wet conditions for sufficently strong warm event forcing. Yet, an increase in rainfall is also observed to occur from coastal British Columbia to northern California for cold SSTs.

The apparent wet signal for both ENSO extremes in the Pacific Northwest is, however, related to different sources. For warm tropical SSTs, the wetness occurs in conjunction with a northward trajectory of mild maritime air masses associated with a deepened and eastward displaced mid-Pacific low. For cold tropical SSTs, the wetness occurs in conjunction with a southward trajectory of cold continental air masses sweeping over the Gulf of Alaska.

b. Analysis of GCM simulations for 1950-94

In this section we seek to determine which features of the observed one-sided regressions are related to true nonlinearties of the atmospheric response, rather than due to sampling artifacts of studying a short instrumental record. Our analysis of the model data will focus on the dynamical aspects of the teleconnections including the behavior of tropical rainfall and 500 mb heights. The regressions of the model data are all derived from the ensemble average of runs for the particular GCM (see Table 1), and are based on simulations for 1950-94.

Table 2 summarizes the amplitude and phase relation of the PNA-sector teleconnection responses to ENSO's extreme phases for the four GCMs and the observations. The root mean square of the warm regressions is consistently higher than it is for the cold regressions in all GCMs. There is some model dependency of those results, for example the CCM and MRF simulations exhibit the greatest amplitude differences for their respective warm and cold SST signals, exceeding those in nature. Importantly, however, the ratio of warm to cold SST rms signals is considerably greater than unity for all the data sets.

Regarding the spatial structure of those teleconnections, the GCMs are once again in agreement that the regression pattern for warm tropical Pacific SSTs is displaced relative to its cold regression counterpart. The extent of the phase shift is, however, quite variable among the models. Both CCM and EC simulations have a -0.8 spatial anomaly correlation between their respective warm and cold regressions, whereas the GFDL model, though having the weakest rms signal, has the greatest phase difference. As will be evident in the following figures, the sense of the phase shift in all four GCMs is to have the warm regression pattern displaced eastward of its cold regression counterpart, in qualitative agreement with observations.

Figures 7-10 display the maps of the one-sided regression responses for 500 mb height, whose salient statistics were summarized in Table 2. Superimposed on these figures are the one-sided rainfall regressions in a fashion equivalent to the observational analysis in Fig. 4. The nonlinearity in tropical Pacific rainfall responses is especially prominent for all the GCMs, a result that reproduces the behavior diagnosed in nature. Note in particular that all the models have a stronger rainfall sensitivity with respect to warm SSTs compared to cold SSTs. As with the midlatitude teleconnections, the equatorial rainfall responses are phase shifted such that the maximum rainfall sensitivity is displaced toward the climatological cold tongue for El Niño conditions, but is displaced toward the climatolgical warm pool for La Niña conditions.

The robustness of the nonlinear rainfall responses among the models is perhaps surprising given the diversity of convective parameterizations used. Such results would imply a particularly strong controlling effect of the equatorial SSTs on the local rainfall, and the longitudinal phase shift of rainfall responses with respect to ENSO extremes emphasizes the importance of total, rather than anomalous sea surface temperatures.

In light of these results, it is useful to examine more thoroughly the relation between ENSO's phase and the atmospheric response, and to determine in particular those SST scenarios under which a nonlinear climate response emerges. Figure 10 presents a scatter plot that compares the amplitude of the EOF1 SST principal component to the GCM rainfall anomalies. The rainfall has been averaged over the region (2N-2S, 150E-150W) for cold SST states, and (2N-2S, 180-120W) for warm SST states. These sample the centers of maximum rainfall suppression (enhancement) during cold (warm) events, and each plotted point corresponds to a winter season during 1950-94. The anomalies are based on the ensemble of each GCM, and these are further averaged across the four GCMs in order to produce an ensemble of 38 model realizations. Scatter plots were also constructed for each

GCM ensemble separately, and those results (not shown) are individually quite similar to Fig. 10.

The scatter diagram reiterates our earlier result of a stronger rate of change in the atmospheric response with respect to warm tropical Pacific SST forcing compared to cold SST forcing. Indeed, the precipitation signal tends to saturate at modestly cold SST anomalies, whereas it increases quasi-linearly with increasing warm SST anomalies. A tapering off of the increase for the extreme 1982/83 warm event is partly an artifact of the region selected for averaging, and large positive rainfall anomalies extended considerably east of the area-averaging region. Overall, an appreciable nonlinearity in rainfall responses emerges only for SST forcings in exceedence of about one standard deviation of the interannual SST variation.

An identical analysis has been performed of the simulated 500 mb height response for a region over the North Pacific. The heights are averaged over the domain (40-45N, 160-150W) where both warm and cold event circulation composites have large signals in the GCMs. Note especially that the functional relationship between the midlatitude height and SST is very similar to that between equatorial rainfall and SST. The similarity between Figs. 11 and 12 is not entirely surprising since the tropical diabatic heating is known to be an important process maintaining the ENSO teleconnections (e.g., Kok and Opsteegh 1985, Held et al 1989). The suggestion is that whereas the midlatitude teleconnections respond nonlinearly with respect to extreme ENSO SST states, they may respond linearly with respect to the ENSO tropical precipitation anomalies.

Despite the large ensemble averages from which these scatter diagrams are composed, it is interesting to note the outliers from the overall linear regression fits that have been emphasized so far. Some of this scatter can be attributed to the fact that the leading EOF of tropical Pacific SSTs fails to account for half of the variance of interannual SST anomalies, and is thus only a gross metric of ENSO itself. Note, for example, the large negative rainfall anomaly of -3 mm/day that occurs in conjunction with a small positive (warm) projection of EOF1 (Fig. 11). In fact, that point corresponds to 1983/84 during which a narrow tongue of anomalously cold water had emerged on the equator, while anomalously warm SSTs existed throughout the off-equatorial tropics related to the remnants on the 1982/83 El Nino. The GCMS are clearly sensitive to the equatorial SSTs in this case, whereas the EOF1 index is reflecting the expanse of off-equatorial warmth which has less bearing on the equatorial rainfall response per se.

Another aspect of the models' behavior as revealed in the scatter plots is less explicable, however. The 500 mb height responses during the three extreme cold SST events are actually weaker than those occurring for much more modest cold SSTs (Fig. 12). A possible explanation may be that SSTs outside of the tropical Pacific are playing a role in these extreme cases. Results to be presented in section 3c, however, reveal a similar behavior in GCM simulations that use only tropical Pacific SSTs.

c. Analysis of GCM simulations forced by idealized SSTs

It is evident from section 3b that nonlinearity in the amplitude of both tropical and extratropical responses is a function of the SST forcing itself. In fact, the response amplitudes scale quasi-linearly for tropical Pacific SST forcings within roughly one standard deviation of the internannual SST variability, and it is principally for the stronger ENSO events that one can discriminate a robust nonlinearity. These inferences have been drawn, however, from GCMs forced with global SSTs, and the fact is that no warm and cold ENSO event combination exists during 1950-94 that exhibit exact opposite SST states.

In order to overcome some of the limitations of the experimental design and analysis methods of section 3b, additional experiments are analyzed in which the GCM is forced with idealized SSTs. These are confined to the tropical Pacific only, and exactly equal but opposite SST anomalies are imposed. As described in section 2, we have performed such runs using the MRF climate model in which the spatial pattern of the SST forcing is given by the leading EOF of tropical Pacific SST (Fig. 1). Monthly tropical Pacific SSTs from 1963-89 are projected onto the EOF1 eigenvector, and these are the only boundary forcing for a nine member ensemble. A parallel nine member ensemble is performed that use the reversed sign of the principal component time series of EOF1. Thus, each observed warm (cold) tropical Pacific SST event during 1963-89 has a counterfactual cold (warm) event twin.

We use simple composite techniques to analyze these new model data, and examine the PNA-sector climate anomalies for warm and cold events separately. In light of some results in section 3b, we form composites separately for weak and strong SST anomalies. The weak SST composites are constructed by averaging winter seasons in which the amplitude of EOF1 is between 0.5-1.0 standardized departure, yielding 5 warm and 5 cold SST cases during 1963-89. In addition, the GCM runs using a reversed phase of the EOF1 are included, and the final composites are based upon an average of effectively 10 warm and 10 cold cases. Note that the composite SST anomalies are now exactly equal, but opposite.

As shown in Fig. 13 for the situation of weak SST forcing, both warm and cold height signals are comparable in amplitude, and there is little indication of a spatial phase shift in the two teleconnection patterns. Likewise, the tropical Pacific rainfall anomalies are also almost equal but opposite, with the largest anomalies located near the equatorial dateline. Thus, to first order, the atmosphere responds linearly to the opposite phases of ENSO for modest amplitudes of the tropical Pacific SST forcing.

The strong SST composites are presented in Fig. 14, and these have been constructed by selecting 1982/83 and 1973/74, the years of maximum warm and cold tropical Pacific SST forcing during the 1963-89 period, respectively. These cases correspond to roughly 2-3 standardized departures of the EOF1 index (see Fig. 1). We once again comingle the simulations based on positive and negative projections of the EOF1 phase. The resulting ensembles have exactly equal but opposite SST anomalies, and each is based on an 18 run average. An appreciable amplitude nonlinearity in tropical and extratropical responses in now clearly seen for these extreme ENSO states, and to accomodate the strength of the atmospheric responses the contour interval in Fig. 14 is now double that in Fig. 13.

Composite height responses over both the tropical and the extratropical Pacific are double for the warm relative to the cold SST forcing, entirely consistent with the regression analyses of section 3b. A similar scaling occurs for the tropical rainfall responses to these extreme ENSO events. Note also the longitudinal phase shift between the warm and cold event responses, and both the tropical Pacific rainfall anomalies and the teleconnection patterns are displaced eastward during extreme warm relative to extreme cold SST forcing.

Finally, is is worthwhile intercomparing the simulated responses for weak versus strong SST forcing. Note especially that the extratropical height signal for cold tropical Pacific SST states is by-and-large insensitive to the amplitude of those SSTs. Indeed, the full impact of cold event forcing on the North American sector can be as readily realized with a 0.5-1.0 standardized SST departure as with a 2-3 standardized departure. The situation for warm events is different in this regard, and the extratropical signal increases quasi-linearly with the amplitude of the SST forcing as previously noted in Fig. 12.

4. Summary

The common practice of studying ENSO climate impacts by constructing the warm minus cold event difference presupposes that the climate signal is strictly linear. In fact, the results presented herein using a suite of atmospheric GCM simulations demonstrate that the assumption, while having merit for weak ENSO states, is poor when applied to the extreme tails of the ENSO distribution.

Two factors motivated our current investigation. One was the desire to explain observational evidence for appreciable asymmetry in the Pacific-North American wintertime climate anomalies during opposite phases of ENSO. In midlatitudes where the ENSO signal competes with strong atmospheric internal variability, it has hitherto been reasonable to dismiss the asymmetries in teleconnections as sampling artifacts. On the other hand, asymmetries in the observed equatorial Pacific rainfall anomalies are manifestly due to the known SST threshold for deep convection, and thus are indicative of nonlinearity. Nonetheless, it has remained unclear which SST states are conducive for such nonlinearity.

The other motivating factor was the desire to reconcile differences among GCM results on nonlinearity of the climate response to ENSO extremes. The idealized SST experiments of Cubash (1985) found no statistically significant midlatitude response to cold event conditions, but a large amplitude response to warm event forcing. Yet, experiments of Hoerling et al. (1997) found similar amplitude midlatitude responses to the warm and cold phases of a simple Rasmusson and Carpenter (1982) SST composite, and experiments for the 1988-89 cold event have also shown appreciable midlatitude responses (e.g., Molteni et al. 1993). The fact is that these studies employ very different SST forcings, and no systematic investigation of nonlinear climate responses to ENSO has been undertaken.

Using the method of one-sided regressions applied both to historical observations and GCM climate simulations of the past half century, we have confirmed the existence and robustness of a nonlinearity in the wintertime climate signals associated with ENSO's opposite phases. It is characterized by a stronger atmospheric sensitivity to warmer tropical Pacific SST states than to colder SST states, and such behavior witnessed in observations was reproduced in four different climate models. The availability of large ensembles of model runs permited a more detailed analysis of the nonlinearity than could be pursued empirically. In particular, it was found that the climate signal for cold SSTs saturates at modest amplitude anomalies such that the atmospheric signals for weak versus strong cold events are virtually identical. In contrast, the atmospheric signal grows quasi- linearly with increasingly warmer tropical Pacific SSTs.

A second characteristic of the nonlinear response is an eastward phase shift of the response for warmer SST states relative to colder SST states. Enhanced equatorial rainfall during stronger warm events expands eastward across the climatological cold tongue, whereas suppressed rainfall during cold events is displaced toward the climatological warm pool. A qualitatively similar phase shift of the midlatitude teleconnections occurs over the PNA region. These shifts, which are apparent in observations, are replicated in each GCM studied, although the degree of spatial displacement between warm and cold event responses is model dependent.

An important conclusion drawn from our GCM analyses is that a necessary condition for nonlinearity in atmospheric responses is that equatorial Pacific SST anomalies exceed one standard deviation of their interannual variability. For SST states at or below a one standardized departure, a virtual symmetry of tropical rainfall and midlatitude circulation patterns was shown to exist For larger amplitude SST forcings, the strength of warm event signals exceeded by two-fold their cold event counterparts, and a spatial phase shift between the signals emerged.

The results have several implications for seasonal atmospheric predictability related to ENSO. First, greater potential predictability exists for extreme warm events compared to extreme cold events. This is so because of a stronger seasonal mean signal during extreme warm events, combined with the fact that the so-called "noise component" of seasonal variability is otherwise independent of SST (Zwiers et al. 1999; Kumar et al. 1999). Second, seasonal skill during extreme warm events could be appreciably sacrificed if based on linear statistical tools. It was rececntly shown that US surface temperature hindcasts for the 1997-98 El Nino winter were more skillfull when using one-sided warm event regressions as the tool, rather than using the linear signal derived from the entire historical record (van den Dool et al. 1998). Furthermore, van den Dool et al. also found that the hindcasts for that El Nino winter were severely degraded when using the reversed sign of the one-sided cold event regression signal.

Lest the case for predictive improvements related to incorporating the nonlinearity of ENSO impacts be overstated, the very requirement for extreme ENSO states to realize such nonlinearity clearly restricts the scope of applications. Likewise, the nonlinearity in midlatitude ENSO signals constitutes a small fraction of the observed inter-ENSO variation of seasonal anomalies, whose principal cause is atmospheric internal varability (e.g., Kumar and Hoerling 1997). Nonetheless, having established the existence for such nonlinearity and the SST conditions under which it emerges has implications for the SST forecast problem itself.

An important question continues to be the error tolerances of the tropical Pacific SST forecasts, both with respect to the spatial pattern of the SST anomalies and the root mean square amplitude error. The studies of Kumar and Hoerling (1997) and Hoerling and Kumar (1997) are especially germaine to the former problem, whereas the results of the current study are most relevant to the tolerance to rms errors. In particular, the saturation of atmospheric responses for modestly cold SST anomalies implies a large tolerance to errors in predicting ENSO's cold phase. On the other hand, the linear increase of atmospheric responses with respect to warmer SSTs places a greater demand to forecast those SST amplitudes accurately. It is the strength of the warm event forcing, through its controlling effect on the strength of the climate signal, that will greatly effect the categorical probabilities, for example of US wintertime surface temperatures. Furthermore, the strength of warm SST forcing will also select somewhat different spatial patterns of the climate signals, quite distinct from the situation for cold event forcing. Specifically, the spatial pattern of cold event responses is insensitive to SST ampltiude, whereas that pattern shifts eastward for progressively warmer SST states.

It is thus with more than acedemic interest that we recognize the existence of predictable patterns beyond the gross linear ENSO signal, and that seasonal forecasts may be improved by incorporating such higher order effects, at least in a few cases. Our focus has been on the wintertime signals over the Pacific-North American region, but it will be fruitful to apply this analysis to other seasons with a particular focus on the SST dependency of tropical monsoon impacts. In this regard, the results of Kumar et al. (1999) regarding the relationship between the Indian summer monsoon and ENSO may be relevant in that their empirical results imply that relationship breaks down during periods of strong east equatorial Pacific warming. The ability to capitialize on these sensitivies in a predictive mode will of course hinge on our predictive capacity for the SSTs. On this matter, Barnston et al. (1999) recently appraised the predictive SST skill of various models during the 1997-98 extreme El Niño and concluded that no model predicted the strength of the warming, at least not until the event was already observed to be very strong. Nonetheless, significant progress has been made in a short period, especially when viewed in the context of recent history in which a comparably strong 1982-83 El Niño event went virtually unpredicted.

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